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Deposition processes for airborne pollutants

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Summary

This is an introductory account of how airborne pollution is ultimately lost from the atmosphere to the Earth's surface. It describes the three main processes: dry deposition, wet deposition and occult deposition, and how they may be measured or estimated. With the problem of 'acid rain' in mind, it assesses what fraction each process makes to the total sulphur deposition in the United Kingdom.

1. Wet deposition

Non-reactive gases are not readily removed from the atmosphere. With this exclusion, precipitation tends to be very efficient at removing most pollutants. Moderate rain will often remove more in one hour than dry deposition will over two or more days. Of course, rain is a very sporadic event. In the lowlands of England it rains only about 3–4% of the time, and even in wet areas like north-west Scotland it rains only some 8–10% of the time. However, as we shall see later, wet deposition is comparable in importance to dry deposition and in the United Kingdom is typically responsible for 70–80% of the total deposition of sulphate, for example.

Pollution is affected by precipitation to a degree which depends on its height in the atmosphere: if the pollution is within the stratosphere it is unaffected since precipitation does not occur there. A typical residence time for pollution within the stratosphere is about 1 year. If the pollution is within the 'free' troposphere (above the boundary layer) it may or may not be affected by any precipitation according to whether or not it is drawn into the cloud. Typical residence times in the free troposphere are thought to be of the order of 10–20 days. If, however, it is within the boundary layer it is subject to dry and wet deposition. A typical residence time for European pollution is 3–4 days.

Pollution comes in a variety of forms:

(a) As particles. One of the most important parameters here is the size of the particle. If it is above about 20 μm (where 1 μm is one thousandth part of a millimetre) then it has a significant fall velocity (see section 2). Smaller particles cannot rely on fallout unless over time they tend to stick together to form progressively larger particles which eventually have a significant fall velocity. This is fairly common and is the main loss route in very arid regions. In most areas, however, small particles are either lost in rain or are carried by turbulence to the ground where dry deposition can take place. The latter, however, tends to be very inefficient for sub-micron particles and they have to rely on wet deposition.

(b) As free gases. Reactive free gases may be adsorbed on to or into cloud droplets, raindrops or snowflakes at a rate which depends on relative gas vapour pressures in the air and in the droplets, sometimes on the pH of the droplets, on the surface area of the drops or flakes, and on how reactive they are.

(c) As captured gases. Gases can be adsorbed on to other particles which are in turn incorporated in to rain or snow. Even some very small particles can become attached to other larger particles of a different kind.

Particles can get into precipitation elements either (a) by being captured, that is, swept out, as the element falls through the air, or (b) by becoming the cloud

condensation nucleus for a developing cloud droplet which subsequently turns into a precipitation element by coalescence. This coalescence may take place simply by random turbulent collisions, or by differences in electrical charge. Particles can also get into cloud droplets as a result of so-called phoretic effects. These are processes by which very small particles and cloud droplets may be brought together by forces acting on the particles pushing them towards the droplet. These forces depend ultimately on the gradient or flux of, for example, temperature, water vapour, electrical charge, or radiation in the vicinity of the droplet. Finally, particles can enter droplets by Brownian capture: when the particles are sufficiently small ($\ll 0.1\ \mu$m) that they are moved by molecular bombardment and can collide with, and be captured by, larger cloud droplets.

Some particles are hydrophobic, that is they are repulsed by water. Obviously they are not good candidates to become cloud condensation nuclei. PCBs are good examples of hydrophobic particles. They can, however, become attached to hygroscopic particles which are particles that tend to attract and hold water molecules to themselves without necessarily dissolving in the resulting liquid water. When this happens, they can be more readily rained out.

Chemical changes can also affect a pollutant's ability to be wet deposited. One of the best known examples of this is the reactive gas, sulphur dioxide, which whilst it is to some extent absorbed directly into raindrops and reaches the ground as sulphur dioxide, can undergo oxidation to sulphate (which is a fine aerosol) either before or after being incorporated into rain. This increases the efficiency of the removal process. The oxidation rate depends not only on the availability of suitable oxidants (like ozone) but also on the relative humidity (or the amount of cloud, see Fig. 1).

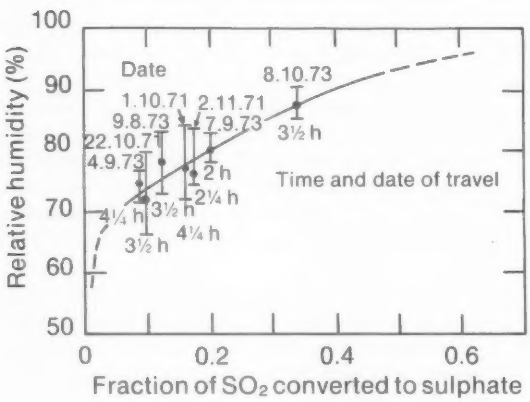

Figure 1. The inferred fraction of sulphur dioxide converted to sulphate as a function of relative humidity. Measurements were made along flight paths off the east coast of the United Kingdom. (Smith and Hunt (1978)).

1.1 Wash-out

From the above discussion, it is clear that pollutants may be removed from the air by precipitation both in cloud and below cloud. Removal in-cloud is generally called rain-out (even though the precipitation may be in the form of snow), whereas below-cloud removal is usually called wash-out. Rain-out is more efficient than wash-out if the concentration is the same in and below cloud.

Considering wash-out first, and following the discussion given by Lyons and Scott (1990) a falling droplet sweeps out an air volume, and its path will intersect that of some of the pollutant particles in this volume. This fraction of particles in the droplet's path is known as the target efficiency, T, and the fraction of these contacting particles that actually stick is the retention efficiency, r. Hence the actual collection efficiency E is

$$E = rT.$$

Very little is known of r and it is usually taken as 1 so that the wash-out coefficient, Λ, is given by

$$\Lambda = \int FTAd(D)$$

where D is the diameter of the falling drop, A is the cross-sectional area of the drop, and F is the number of drops in interval $d(D)$ per unit volume, times their fall-speed.

If the wash-out process occurs uniformly over the plume volume, then the concentration in the air decreases in time t due to wash-out alone according to

$$C_a(t) = C_a(0) \exp(-\Lambda t).$$

This method is, strictly speaking, only applicable to particles of a single size and to highly reactive gases which are irreversibly captured by the precipitation. Normally the method can only be applied in an empirical sense if a range of particle diameters is involved.

The resulting concentration in the rain may be expressed in terms of a related wash-out ratio ω:

$$C_r = \omega C_a$$

where C_r is the resulting concentration in the rain. If C_a and C_r are expressed in g m^{-3} then ω is usually of the order of 10^5. If, however, C_r is expressed in g l^{-1} then ω is of the order of 10^2. Note that ω depends on the total depth of the polluted column through which the rain falls. The values quoted refer to a depth of the order of 1000 m.

1.2 Wet deposition

The resulting wet deposition D_w is obtained by identifying the concentration in the rain in terms of g l^{-1} multiplied by the rainfall R in millimetres as the

deposition in terms of g m^{-2}.

$$D_w = C_r \times R = \omega C_a R.$$

The wash-out ratio can also be used to define a wet deposition velocity by analogy to the dry deposition velocity (see later).

$$v_w = \frac{dD_w}{dt} / C_a = \omega \frac{dR}{dt}$$

where t is here time, and the units of v_w are here in m s^{-1}. With light rain falling at a rate of 1 mm h^{-1} and a value of ω of 100, v_w is about 3 cm s^{-1}, which is of the order of a 1000 times greater than the corresponding dry deposition velocity.

Wash-out ratios for snow tend to be several-fold higher than for rain although R for snow is usually less than for rain due to the lower temperatures involved.

Λ is related to ω through the relation:

$$\Lambda = \frac{\omega}{h} \frac{dR}{dt}$$

where h is the depth of the polluted layer through which the rain is falling, R is in mm, and t, the time, is in seconds. Assuming $\omega = 100$, $h = 1000$ metres, then Λ is approximately $3 \times 10^{-5} R'$, where R' is the rainfall rate in mm h^{-1}. Jones (1983) suggests a value of Λ between 3×10^{-5} and 3×10^{-4}, consistent with our estimate when rain-out is artificially included and when we allow for heavier rainfall rates.

Note that we have assumed a linear relation between Λ and R. However, some theories and experimental evidence have pointed to a dependence:

$$\Lambda = aR'^b$$

where b is not 1 but lies in the range 0.7 to 0.9.

1.3 Rain-out

Rain-out is, as previously stated, generally even more efficient than wash-out if the pollution gets into the precipitating cloud. The relationships can be made exactly analogous to the relationships for wash-out. However ω is no longer a function of the depth through which the rain falls. Its magnitude is frequently 5–10 times greater than for wash-out. For reactive gases and hygroscopic particulates a value around 500 is typical. (The deposition of caesium-137 on the United Kingdom from the cloud of debris from the Chernobyl reactor accident implied a total ω of about 650 (Smith and Clark 1989).) The corresponding value of Λ is $1.5 \times 10^{-4} R'$, where R' is the rainfall rate in mm h^{-1}.

Note that if the equation for the deposition in terms of C_a and R uses mutually consistent units, so that R is given in metres, then $\omega = 5 \times 10^5$.

1.4 The measurement of wet deposition

By comparison with dry deposition, wet deposition is fairly easy to assess, although it certainly has its pitfalls. These pitfalls include:

(a) the difficulty of accurately measuring rainfall and snowfall,

(b) the variability of precipitation between sites separated by relatively short distances, especially in topographically complex terrain (like mountainous areas),

(c) with some pollutants the difficulty in sampling sufficient material to analyse,

(d) contamination or degradation of the collected rainwater before analysis takes place, and

(e) the additional effects of dry deposition whose representativeness in terms of deposition on surrounding natural surfaces is largely unquantified.

Generally it is best to use 'wet-only collectors' that are electrically controlled to open only when precipitation is falling, and the collected sample is kept at sufficiently low temperature to minimize degradation. It also helps to collect the samples at frequent intervals (e.g. daily) and analyse them as quickly as possible. Some research collectors can analyse the rainwater internally as soon as the amount in the collecting vessel reaches some prescribed amount. 'Bulk' collectors are open all the time and sacrifice some accuracy for the cheapness of the equipment. It is usually best with both types of collector to measure the absolute amount of rainfall with a properly exposed approved type of rain-gauge.

1.5 The seeder-feeder effect

In mountainous terrain a substantial proportion of the rainfall originates from the scavenging of cloud droplets in cloud capping the mountain tops, called 'feeder cloud', by rain falling from higher frontal cloud, called 'seeder cloud' (see Fig. 2 and Browning *et al.*

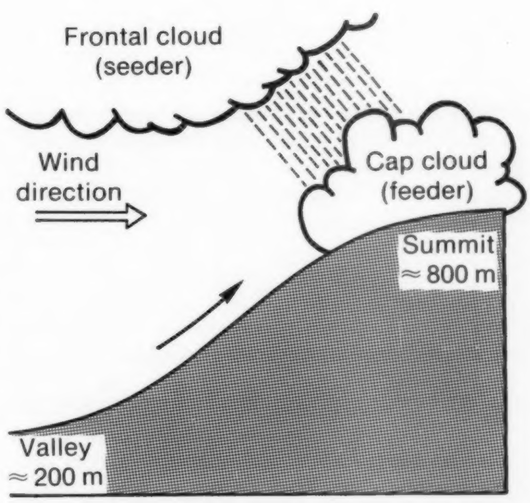

Figure 2. Seeder-feeder mechanism for enhanced rainfall concentrations of major ions in rain from Browning *et al.* (1974).

(1974)). The figure schematically shows the situation at Great Dun Fell which has been investigated by several research groups (see *Acid Deposition in the United Kingdom 1986–1988* Third Report of the UK Review Group on Acid Rain, Dept of the Environment, London). Two factors are important. The first is that the rainfall is enhanced by altitude. The enhancement depends on a number of factors including the details of the orography in the area. At Great Dun Fell the rainfall at the summit is roughly twice that at the upwind valley bottom. Fig. 3 shows a very broad-brush relationship between rainfall, normalized by an estimate of the equivalent mean-sea-level rainfall in the same area, as a function of the height of the mountain/hill tops. It should not be used as an exact relationship. It is simply inferred from comparing the average annual rainfall map for the United Kingdom with a physical map showing height contours. The second factor of importance is that since the feeder-cloud water often has concentrations of ions that are several-fold higher than those in the seeder-cloud, the concentrations of these ions in the rain reaching the ground over the hills is often much higher than in corresponding rain over lower ground. At Great Dun Fell the enhancement in concentration is about 2.5 so that the deposition, the product of concentration and rainfall, is enhanced five-fold at the summit. More generally, where simultaneous measurements have been made of orographic cloud and rain composition, the concentrations of major ions (like sulphate) in the cloud usually exceed those in rain by a factor ranging from 1.5 to 8.

The *Acid Deposition in the United Kingdom 1986–1988* Report, referred to above, has estimated the percentage enhancement of wet-deposited nitrate due to the seeder-feeder mechanism on a 20 km × 20 km grid. The seeder rainfall was estimated by linear interpolation between west coast rainfall and east coast rainfall. Actual rainfall in any grid square was inferred from the average annual rainfall map, and the excess over the inferred seeder rainfall was ascribed to the additional scavenging of feeder-cloud water whose concentrations of major ions were assumed to exceed those in seeder rain by a factor of two. Fig. 4 shows the results. It indicates that fairly substantial areas in the north and west may have 60% or more enhancement of nitrate deposition due to orographic effects.

Choularton in Choularton *et al.* (1988) and in his contribution to the report cited above has expressed these enhancements in terms of some of the parameters described earlier. He suggests that wherever the weather radar suggests that cloud is in contact with the surface (excluding radiation fogs) occult deposition (see later) can be modelled with an enhanced deposition velocity of $10\ m\ s^{-1}$ for the pollutants believed to be within the

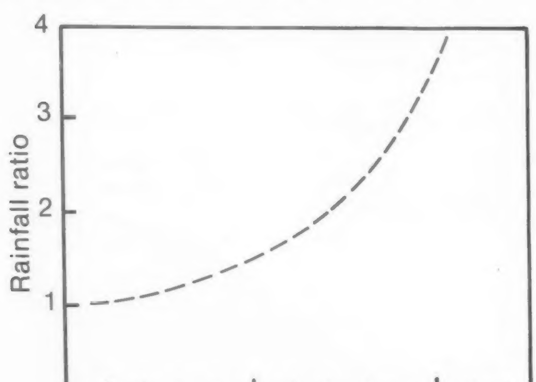

Figure 3. A rough guide to how rainfall, normalized by the equivalent sea-level rainfall in the same area, increases with terrain altitude.

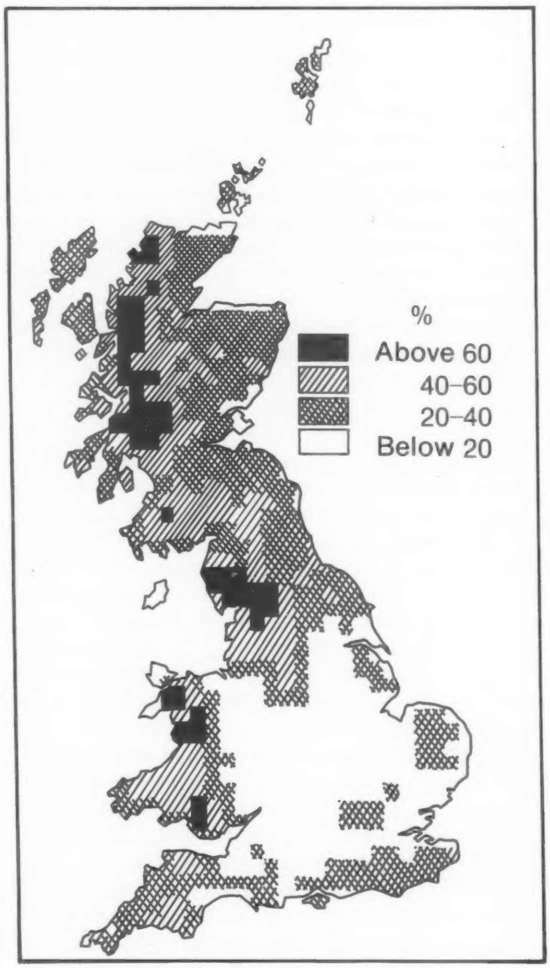

Figure 4. The estimated percentage enhancement in rainfall due to the seeder-feeder mechanism inferred by the method described in the text, due to Fowler *et al.* (1990).

cloud droplets, and that if the seeder-feeder mechanism is operative an enhanced scavenging coefficient should be used:

$$\Lambda = 3.4 \times 10^{-4} R'^{0.79}$$

where, again, R' is the rainfall rate in mm h^{-1}.

1.6 An important caveat

It should be stressed that this treatment of wet deposition has glossed over a great number of issues and complexities. Many of these are discussed by Slinn (1984).

2. Dry deposition

2.1 Introduction

The direct take-up of airborne gaseous or particulate material present in the lowest layers of the atmosphere to the surface or to vegetation by processes of absorption, impaction or sedimentation is called dry deposition, to distinguish it from deposition by precipitation, even though at times dew, cloud droplets or surface water may be involved.

Dry deposition on natural surfaces is very difficult to measure directly, certainly in any routine operational manner. Some alternative methods will be outlined later. The supposition is usually made that the rate of deposition, or flux, F, is proportional to the concentration in the air just above the surface at height z:

$$F = v_d\, C(z).$$

The coefficient v_d has the dimensions of a velocity and is therefore called the deposition velocity, and is a slowly varying function of height z. Normally it may be assumed that v_d is independent of the magnitude of $C(z)$ although it must depend on the physical and chemical characteristics of the species involved. Sometimes, however, the uptake of the species at the surface may be affected by how much of the species is already deposited. A good example of this is the reduction in the deposition rate of an acidic species on to a damp surface as the pH of the water film on the surface decreases as more of the species (and other acidic species) accumulate in the film.

If $C(z)$ is measured a metre or so above the ground then the stability of the air will influence how rapidly turbulent transfer can carry species down to the surface. In stable conditions, this transfer may become very slow as atmospheric turbulence is suppressed. The magnitude of $C(z)$ (for any specified emission rate) may also respond to changes in stability. Consequently v_d and $C(z)$ may both vary with stability in some systematic way. Consequently $\overline{v_d\, C(z)}$ averaged over a variety of stability conditions may be significantly different from $\overline{v_d}\ \overline{C(z)}$ (where the individual averaging is over the same ensemble of conditions). Similarly the ground may be both a potential source and sink of some species like ammonia (for example) and whether source or sink at any one time may depend on many factors likely to include the time of day, the time of year, temperature and other factors affecting relevant soil and vegetation processes. Here again v_d may take a range of possible values which correlate with $C(z)$.

In most real-life scenarios, surface conditions are not totally homogeneous so that v_d may change from point to point. The question of how to average v_d under these circumstances has not been entirely resolved, but a plausible suggestion is that v_d may be averaged in a way related to the way the surface roughness is averaged over heterogeneous terrain. If this is so then the average v_d is rather strongly influenced by a few 'hot-spots' of high-deposition rate and also might be a significant function of wind direction at any point.

2.2 Expression of deposition velocity in terms of resistance

Over horizontally uniform surfaces, the downward vertical flux may be expressed in terms of the eddy diffusivity $K(z)$:

$$F = K(z)\,\frac{dC}{dz}.$$

If F is virtually constant with height in the first 10 metres or so, then

$$C(z_2) - C(z_1) = \int_{z_1}^{z_2} \frac{F}{K(z)}\,dz \equiv F\int_{z_1}^{z_2} \frac{dz}{K(z)}$$

which is an analogous to Ohm's Law in electricity. The integral can therefore be compared to a resistance r. If $z_1 = 0$ and $C(0) = 0$, then

$$v_d^{-1} = \int_{0}^{z_2} \frac{dz}{K(z)} = r.$$

However, $K(z)$ is not defined in any simple way as $z \rightarrow 0$ corresponding to the actual physical surface. Consequently the resistance r is normally subdivided into an aerodynamic part plus parts appropriate to close to the surface:

$$r = \frac{1}{v_d} = r_a + r_b + r_s$$

where

r_a = the resistance to transport through the turbulent zone,

r_b = the resistance through the laminar sublayer very close to the physical surface,

r_s = surface resistance which will depend on one or more of surface moisture, pH, the reactivity of the species, the physical properties of the species and the status of the stomata on any vegetation.

By analogy with the flux of momentum

$$r_a = \frac{u(z)}{u_*^2} .$$

The sublayer resistance r_b has been expressed by Chamberlain (1960) as

$$r_b = \frac{1}{B u_*}$$

where B is a sublayer Stanton number. Chamberlain showed that B decreases slowly with increasing u_* but varies very little with surface roughness z_o. Typically $B \approx 0.125$ for gases.

The surface resistance is quite variable and the hardest of the three to estimate accurately. Rather typical values for a short grass field are as follows:

(a) wet conditions, $r_s \approx 0$,

(b) dry daytime conditions, $r_s \approx 50$–100, and

(c) dry night-time conditions, $r_s \approx 300$–1000.

Consequently semi-empirical values of v_d can be evaluated. For example, if $z_o = 0.1$ as it might be over growing wheat, and the stability is near neutral, then the values of v_d in Table I apply for a reactive gas.

Table I. Typical values of v_d over growing wheat when the wind speed and the concentration of the depositing reactive gas are measured at 10 metres in neutral stability conditions

Conditions	$U_{10} = 2\ \mathrm{m\,s^{-1}}$	$U_{10} = 8\ \mathrm{m\,s^{-1}}$
Wet	$0.89\ \mathrm{cm\,s^{-1}}$	$3.56\ \mathrm{cm\,s^{-1}}$
Day dry	$0.53\ \mathrm{cm\,s^{-1}}$	$0.97\ \mathrm{cm\,s^{-1}}$
Night dry	$0.16\ \mathrm{cm\,s^{-1}}$	$0.19\ \mathrm{cm\,s^{-1}}$

The overall likely average of these numbers when averaged over a year are in reasonable accord with Fowler *et al.*'s (1991) estimate for the whole of the United Kingdom with all its different surfaces. His estimate for sulphur dioxide is $0.24\ \mathrm{cm\,s^{-1}}$. Fowler divided the United Kingdom into 20×20 km grid squares and used up to five land-use categories in each square using a classification method developed at the UK Institute of Terrestrial Ecology. The atmospheric resistances ($r_a + r_b$) were calculated from the z_o of the vegetation and average wind speeds assuming a logarithmic wind profile in the surface boundary layer. For each vegetation type a seasonal pattern of leaf area was defined, and the physiological processes controlling gas uptake by stomata were modelled using 'big leaf' methods for a range of vegetative species. The resulting v_d show a marked diurnal and annual variation, as exampled by the variation over a pine forest as seen in Fig. 5. Fowler notes the pronounced effect of day length on stomatal opening and hence on the deposition velocity. The very small December increase in v_d is due also to the temperatures then being close to the threshold for stomatal opening. In midsummer the value of v_d dips slightly at midday due to typical water stress on the trees at that time.

2.3 Dry deposition of particles

The deposition of particles to surfaces depends very much on their ability to cross the laminar sublayer. Large particles whose diameters exceed 10 μm are usually heavy enough to fall through by sedimentation or by the inertia of the particles. In contrast very small

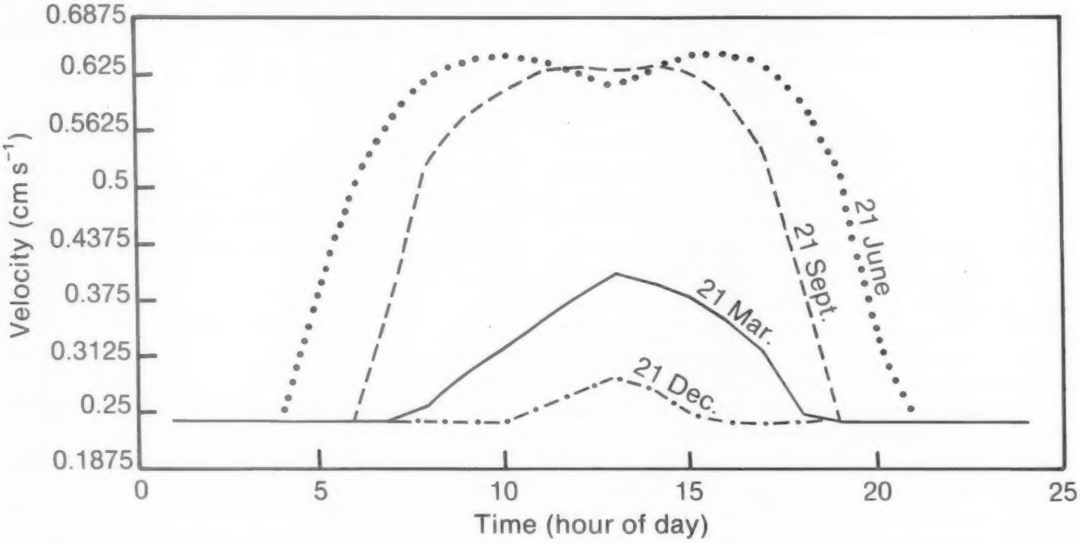

Figure 5. The diurnal cycle in the deposition velocity of sulphur dioxide gas to a UK forest at four different times of the year (Fowler *et al.* (1990)).

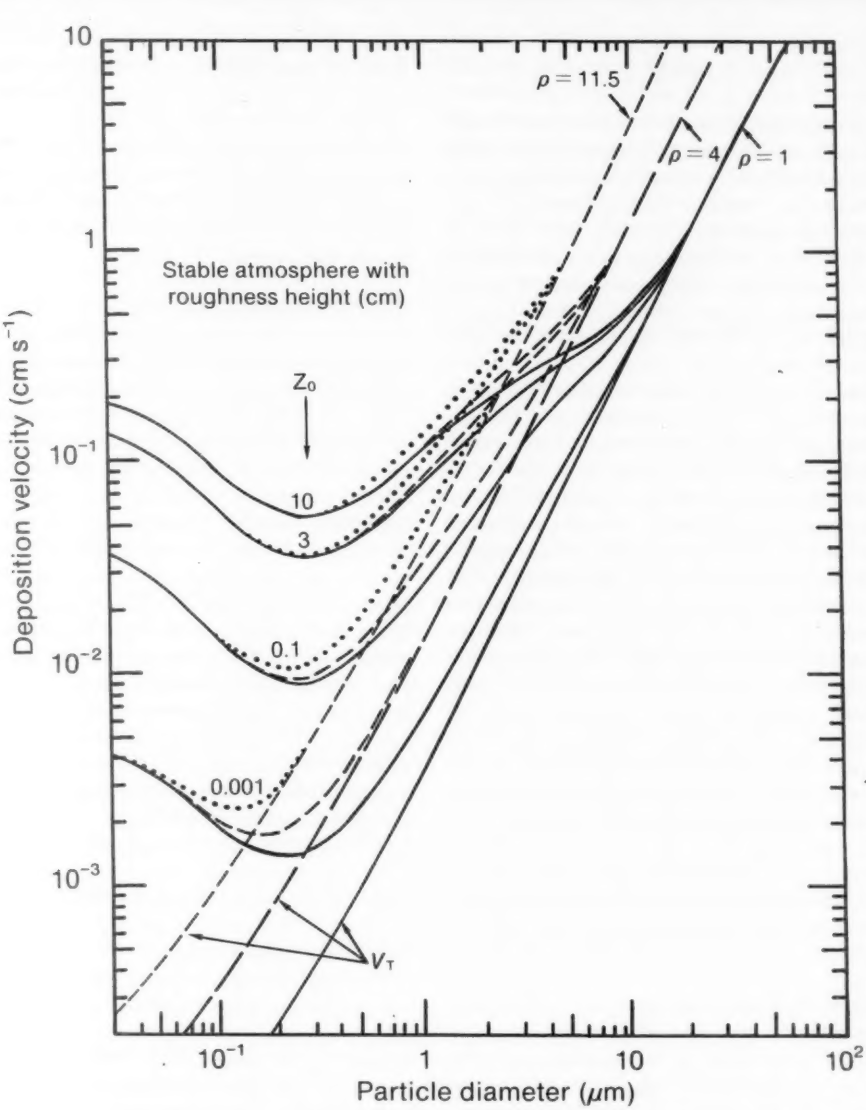

Figure 6. Extrapolation from correlations of wind-tunnel measured deposition velocities at a reference height of 1 m, for particles of density 1, 4 and 11.5 g cm^{-3} (after Sehmel (1980)). V_T represents the terminal settling velocity. The diagram comes from Nicholson (1988).

particles whose diameters are less than 0.1 μm are quite responsive to Brownian motions and can diffuse across the layer. Intermediate sized particles however tend to be too large to be affected by Brownian motions and too small to have sufficient inertia or weight to penetrate the layer. These sized particles therefore display minimum values of v_d within the range of diameter as can be seen in Fig. 6 which comes from a review paper by Nicholson (1988).

2.4 The measurement of estimation of dry deposition velocity

Various methods have been devised for estimating the dry deposition velocities of various species in various field conditions (see Hicks (1986) for a fuller description). On the whole it seems these methods tend to be more successful overall for gases than for particulates although the reasons for this are not fully understood. The different methods can be classified as follows: micrometeorological methods (profile methods and correlation methods), collection on to surrogate surfaces, collection on to natural surfaces, and budget techniques.

2.5 Profile method

This method depends on a direct analogy between the downward flux of the depositing species and the flux of heat and momentum. It uses the flux-gradient relationship

$$F = K(z)\frac{dC}{dz},$$

where $K(z)$ is inferred from the wind profile. In near-neutral conditions $K(z) \approx ku_*z$. This method, as well as the correlation method, involves collecting data over a period of time that is unlikely to exceed at most a few hours. These methods are therefore best suited to studying the effects of environmental conditions on deposition mechanisms. The profile method is only suitable for a limited range of species, mainly the more reactive gases, and is only recommended when other methods cannot be used. A typical error in v_d for a reactive gas, inferred from data collected over 30–60 minutes, is about 0.05 cm s^{-1}. At best it is difficult to measure the small gradients accurately, partly because of the wide range of time-scales involved in the concentration fluctuations. Moreover, it is advisable to use the same intrument to measure the concentrations at the different heights to avoid spurious gradients implied by different instruments with slightly different responses. It is also expedient to sample over a large range of heights within a layer that is nevertheless shallow enough to be contained within the so-called constant-stress layer. It is also important to make these measurements within a period of time when the meteorological conditions are not changing. Finally the site has to have a large uniform upwind fetch so that the profile is not affected by upwind variations in deposition rate and surface roughness and sensible heat flux.

It is not surprising that with all these caveats the method is prone to significant error. The method appears to work best for gases. Particles of a few μm in size are probably the largest that can have their v_d estimated by this approach.

2.6 Correlation methods

Eddy-correlation techniques require the simultaneous measurement of vertical velocity and concentration with sufficient time response over the range of frequencies in which the two are effectively correlated. Unfortunately while there are a number of instruments capable of measuring certain gaseous concentrations with adequate response, there are few instruments available for measuring particle concentrations on the right time-scale (thought to be about half a second). Electrical and optical sensors have been used, as have flame photometric detectors. The method depends on the identity $F = -\overline{wc}$.

It is necessary to eliminate any small residual mean vertical velocity due to a finite sampling period or due to the peculiar position of the site or due to a false alignment of the anemometer. Elimination can be realized by numerical or electronic filtration of the data before the mean product is calculated.

The eddy-correlation technique has provided the most consistent and controversial differences between field estimates of v_d and those estimated from theory and wind-tunnel studies. In particular, values found by the flame photometer detectors seem to be consistently an order of magnitude greater than those found from the latter methods. They also sometimes give unexpected upward fluxes at night, and the overall impression is that further study of this method is required before its findings can be trusted. Differences encountered with particles may be due in some cases to the presence of large particles, although this requires verification in actual field studies.

2.7 Surrogate surfaces

The use of surrogate surfaces appears to be successful although the results strictly only apply to those surrogate materials and not necessarily to natural surfaces in the same situation. Wet/dry bucket samplers are thought to work best for particles with a significant terminal velocity. Results for certain radionuclides appear to agree with known inputs and atmospheric concentration fields. Success with some of the acidic species is less certain because of the difficulty in totally eliminating contamination from wind-blown dust, from insects and from bird droppings. Surrogate collectors may also disturb the airflow and are not representative of the real surface of interest. For example, buckets have been found to over-collect particles with diameters much greater than 1 μm but under-collect sub-micron particles. Consequently it is very important to know the particle size spectrum and to be able to correct for these collection effects to some reasonable extent. Although surrogate surfaces seem best for larger particles, gases have also been collected on special filter-paper surfaces.

Overall the results from surrogate surfaces are comparable to those found by other reliable techniques.

2.8 Natural surfaces

Perhaps the most obvious method for particles is to use natural surfaces. Whilst leaves can be removed and the particles washed off and analysed, it is much harder to deal with creviced bark and some other natural materials. These surfaces are much less convenient in the sense it is harder to wash-off the particles for analysis, and they may be 'contaminated' by wet deposition. Another problem is that vegetation may take up the same species through their roots and these may be partially leached to the leaf surface and affect the results of analysis.

The estimates of v_d obtained from a limited number of samples may also be prone to error because of the wide variety of orientations and exposure of natural surfaces.

Measurements of deposition to snow have been frequently studied. The analysis is relatively easy and is representative of fairly uniform surfaces. Since the deposition rates depend on particle size, the greatest difficulty encountered here is to measure accurately the particle size spectrum in the air itself.

Overall the agreement between results from this method and those from theory and from wind-tunnel experiments is generally poor. In addition to the reasons

already postulated, the effect of the complicated flow pattern around trees and other vegetation could easily have a profound effect.

2.9 Budget methods

This method is perhaps applicable when concerned with transport over very long distances well in excess of 100 km and only when a broad average deposition velocity is required. Two approaches may be used. The first is to optimize the agreement between model predicted air concentrations (using an operational model like the EMEP (European Monitoring and Evaluation Programme) model for acid species over Europe) and observed daily-averaged concentrations measured at many monitoring stations across the region by adjusting the value of the deposition velocity in the model. For example the EMEP model allows for loss of the species from the boundary layer into the free troposphere above, only in a highly parametrized manner, and any systematic error would result in an erroneous v_d.

The second approach is to compare the total downwind flux of the species using an instrumented aircraft to fly at different levels across the plume with either the emission strength at the source or the total flux at some other distance separated sufficiently from the first to yield a measurable difference of flux. This has been done several times by the Meteorological Office and The UK Central Electricity Research Laboratories over the North Sea in which the flux of sulphur dioxide and sulphate originating from a large power station has been studied.

On a smaller scale the same principle can be applied although it is then often necessary to label the particle species, for example by using an isotope of the species which can be readily detected *in situ* by a Geiger counter, for example.

2.10 How important is dry deposition to the long-term total deposition?

On the whole, micron-sized and submicron-sized particles dry deposit relatively slowly with deposition velocities smaller than $0.1\ \mathrm{cm\,s^{-1}}$. However many of them are rather efficiently washed out or rained out in precipitation. Under these circumstances dry deposition only contributes a rather small proportion of the total deposition in Europe. Reactive gases like sulphur dioxide dry deposit more readily. If we take Fowler's estimate of $0.24\ \mathrm{cm\,s^{-1}}$ for sulphur dioxide as valid, then in a location which has an annual average rainfall of, say, 800 mm and assuming the simple wet deposition law discussed in section 1.2:

$$\text{wet deposition} = D_w = \omega C R$$

where C is the average air concentration and R is the annual rainfall in millimetres, and w is a coefficient whose magnitude is around 500. If $C = 10\ \mu\mathrm{g\,m^{-3}}$ then the annual dry deposition is about $0.76\ \mathrm{g\,SO_2\,m^{-2}}$ and the wet deposition would be $2\ \mathrm{g\,SO_2\,m^{-2}}$, on the assumption that in rain the average concentration in the air is only about half the annual average since rain cleans the air so relatively quickly. Dry deposition would then constitute only about 32% of the total. Also since it typically rains only about 3–4% of the time this means that wet deposition is about 50–100 times more efficient than dry deposition when it is actually occurring.

2.11 Effect of dry deposition on long-range transport

Neglecting the spatial variations in the deposition velocity that must occur, the effect on the amount of material still airborne can be represented rather simply by a source-depletion method provided the value of v_d is not too high (i.e. comparable to the vertical turbulent velocities). In the simplest models of long-range transport the easiest way to allow for dry deposition is then by assuming that the pollutant is at all times vertically well-mixed within the boundary layer. The concentration in the air is then a function of the horizontal co-ordinates x and y. Integrating acrosswind $C = C(x)$. The variation with x is then given by:

$$\frac{\partial C}{\partial x} = \frac{1}{u}\frac{\partial C}{\partial t} = -\frac{1}{u}\frac{v_d C}{h},$$

i.e.

$$C = C_o \exp\left(-\frac{v_d x}{uh}\right)$$

where h is the depth of the boundary layer. Typically $h = 600$ m, $u = 8\ \mathrm{m\,s^{-1}}$ and, for sulphur, $v_d = 0.2\ \mathrm{cm\,s^{-1}}$ so that the length-scale, L, is given by:

$$\text{dry deposition length} — L = \frac{uh}{v_d} = \frac{600 \times 8}{0.2 \times 10}\ \mathrm{km},$$

i.e.

$$L = 2400\ \mathrm{km}.$$

Since the flux F of the material to the ground is proportional to $v_d\, C(x)$, differentiating with respect to v_d and putting the result to zero, says that the rate of dry deposition is very insensitive to the exact value of the overall v_d at downwind distances between about 1000 km and 4000 km. The overall decrease in concentration with larger v_d virtually balances the larger v_d in the expression for the dry deposition within the range of distances.

3. Occult deposition

Sometimes cloud or fog is at ground level, particularly on hills in maritime climates. Pollutants within the boundary layer are advected up the hillside and become absorbed into cloud droplets as the moisture in the air condenses or may form the condensation nuclei on which the droplets grow. Soon after the condensation

level the droplets are often large enough so that as they blow across the vegetation they are deposited on to the foliage, carrying the pollutants with them. This form of deposition is called occult (or hidden) deposition because it is not easily assessed from standard meteorological data.

The capture of wind-driven cloud droplets by vegetation has been measured in several field experiments (Gallagher *et al.* 1988, Fowler *et al.* 1991). These experiments show that the process is efficient with equivalent deposition velocities of 1–4 cm s^{-1} over moorland and in excess of 10 cm s^{-1} over forests. Moreover, the relatively large concentration of major ions in cloud water (50–2000 μeq l^{-1}) increases the importance of this pathway in upland areas. Fowler *et al.* (1991) have estimated the average annual occult deposition rate over the United Kingdom. They did this as follows:

(a) The country was divided into 20 × 20 km grid squares.

(b) Cloud-base statistics were interpolated from meteorological records to each of the squares.

(c) Wind velocity data were also interpolated to the squares.

(d) A land-use database was used to infer the proportion of land within each square which is either forest, moorland, grassland or arable crop-land. The altitude of these areas was also obtained.

(e) The data from (d) were used to obtain average z_o, and hence v_d.

(f) Surface cloud was assumed to contain concentrations of major ions a factor of 2 larger than in normal rain, as measured in the UK air pollution 'secondary' network.

(g) Liquid water content of the cloud water was taken as 0.2 g m^{-3}.

Resulting implied occult depositions, averaged over each 20 × 20 km grid square, ranged from very small values in most lowland areas to in excess of 160 mgS m^{-2} a^{-1} for the sulphur ions in some of the mountainous areas of Scotland (in the Cairngorms) where normal dry deposition is only about 40 mgS m^{-2} a^{-1} and wet deposition is about 600 mgS m^{-2} a^{-1}.

Radiation fogs, which occur rather frequently in some maritime lowland areas in winter, contribute relatively little to the annual deposition of major ions because the wind speeds are normally very small under these conditions.

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On the use of numerical probabilities in weather forecasting

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Summary

A concept of risk assessment in weather forecasts expressed as numerical probabilities is considered from the point of view of the customer as well as that of the forecaster. For the customer the information quantifies the risk of specific events occurring or the risk of specific thresholds being exceeded, e.g. temperatures falling below 0 °C during defined periods of time. For the forecaster the concept is an intrinsic one because it allows him to express his forecasts in a language that is ideally suited to his perception of the evolution of a weather system on any occasion. For both customer and forecaster the concept has the immensely powerful attribute that it is amenable to objective evaluation thereby providing a measure of skill and an assessment of the forecaster's value to the customer.

1. Introduction

The notion that weather forecasting is an inexact science is (probably) properly understood by only a small number of enlightened people. Most interpret it to mean that forecasts cannot be relied upon for accuracy — defined in some way. What is not appreciated is the reason why the accuracy can vary and that it is possible for a forecaster to anticipate the likelihood of the forecast being within or outside the range of average accuracy as defined. The key word is 'accuracy' and the definition will almost certainly depend upon what is at stake. For example horticulturists are not too concerned if the temperature falls to +8 °C rather than +11 °C but they would be very concerned if it fell to 0 °C rather than +2 °C. Similarly offshore drillers are not too concerned whether the wind is 5 kn or 30 kn but if it reaches 65 kn rather than 55 kn they are very concerned because the risers are liable to break at such speeds. These examples reveal that the most important consideration for many users of weather forecasts is risk assessment in relation to specific thresholds, rather than categorical forecasts.

The use of probabilities in weather forecasting in the United Kingdom is extremely limited and, until recently, confined entirely to some commercial forecasts. The most extensive use is found in forecasting for aviation, for aerodromes as well as *en route* weather. On the face of it, it is surprising that greater use is not made of risk assessment forecasts by industry and commerce since the concept would appear most suitable for their needs. The main reason for this could be the lack of clear guidelines issued by the forecasters on how to interpret probability forecasts. For example a recent study in the USA, where probabilities are used in public forecasts disseminated on radio and television, showed that people were confused as to whether a 10% chance of rain meant that 10% of the area would have rain or whether there was a 1 in 10 chance of rain affecting anyone. (Despite this apparent ambiguity the public appear to prefer numerical values to words by a 2:1 majority.)

Then there is the tricky question of how the forecast is to be used to make commercial decisions; given a 30% chance of frost the horticulturist can weigh up the risk to his flowers in terms of cost to himself but he also needs to know something about the quality of the forecast service too. In other words is that 30% chance realistic, pessimistic or optimistic. Fortunately probability forecasts are very amenable to verification which helps in making these decisions.

Public interest in risk assessment attached to weather forecasts has increased in the United Kingdom, almost certainly as a consequence of the severe winter storms experienced in early 1990. National radio and television companies are now keen to issue forecasts in this format to the public. It is appropriate therefore to look at the concept of probability forecasting from the practical point of view of the forecaster and the user who has to interpret the information.

2. Qualitative forecasts

Descriptive weather forecasts are most widely used for issue to the public through Press, radio and television, Videotex and TIS (Telephone Information Services), etc. It is a very convenient way of describing weather conditions likely to affect an area over a period of time. Unfortunately, in achieving this objective there is a loss of precision in the forecast. Subjective interpretation by the user of the information is paramount and it is very difficult to verify the accuracy of the forecast. Risk may be mentioned in the forecast, e.g. 'perhaps later' but never quantified.

A descriptive forecast serves a very useful purpose as an overview of the weather situation or as a preamble to a more detailed site-specific forecast, but except in the most simplistic weather situations its utility is distinctly limited. The information content is scanty and the scope for misinterpretation of language is large.

3. Categorical forecasts

Quantitative forecasts of weather elements, e.g. wind, temperature, precipitation, cloud, etc. are issued widely to industry and commerce on a commercial basis. Wind and max./min. temperatures are also issued to the public. Values of the elements are necessarily for fixed sites and usually for fixed times. Interpolation problems are minimized by increasing the number of sites and decreasing the temporal spacing of forecasts at each site through the forecast period. These forecasts are precise and categorical and many commercial customers insert the figures into prediction models to help with their own decision-making.

To the forecaster, categorical values of weather elements represent the best estimate that the man–machine mix can provide. To the user, the information is unequivocal and unambiguous but, in making his own decisions, he still has to make some allowance for forecast error.

Assuming that reliable and accurate observations are available, categorical forecasts are easily verified. The most common parameters derived for this purpose are the mean error and the root-mean-square (r.m.s.) error. Both can be applied to scalar and vector fields. The mean error gives an indication of any bias and the r.m.s. error gives a measure of the spread of forecast values from the actual values obtained. Given a zero bias and normal distribution of data, the user may expect about 84% of all forecasts to lie within one r.m.s. error and 98% of all forecasts to lie within twice the r.m.s. error range of accuracy. Figs 1(a) and 1(b) illustrate typical r.m.s. error distributions with increasing forecast period for surface wind and surface temperature for specific locations. The error values are calculated from a large sample of data. Superimposed on each graph is a specific set of forecasts issued on one occasion for that location. The recipient of the forecast data can assess the implications of the range of error at each forecast time.

Consider Fig. 1(a); in fact the r.m.s. values plotted here are the mean for a cluster of six offshore platforms in the central and northern North Sea and indicate a steady increase of error as the forecast period increases. The forecast values are for one platform in the cluster with a peak wind speed of 30 m s^{-1} at T+96. If the Offshore Installation Manager (OIM) interprets this forecast sensibly he would deduce that a reasonable expectation is for the actual wind to lie in the range 25–35 m s^{-1} but he cannot ignore the possibility that the actual value will lie outside this range. The statistical accuracy can be presented in a slightly more revealing way by looking at the errors with respect to wind speed observed. Fig. 2 shows this for four bands of wind speed and not surprisingly the errors are rather greater for the higher speeds; note too that the mean errors show a small positive bias at low speeds and a definite negative bias at high speeds. The most practical use that the OIM can make of this information is to recognize that if a stormy spell of weather is expected then the forecasts are likely to underestimate the severity at some stage.

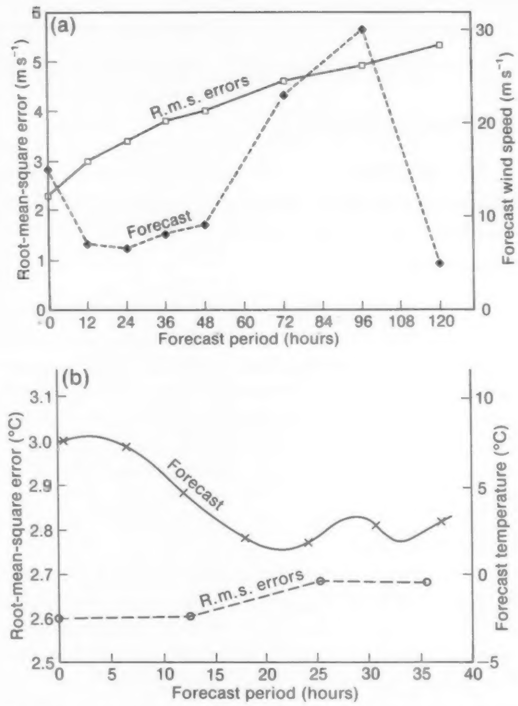

Figure 1. Typical root-mean-square errors with respect to forecast period, and a specific forecast of site values for (a) wind speed and (b) temperature.

4. Causes of variation in accuracy

Weather forecasts are largely based upon Numerical Weather Prediction (NWP) models nowadays, with the (human) forecaster having the role of fine-tuning the product and adding value to it. There are two fundamental reasons why NWP models do not yield perfect forecasts: firstly there is the limitation of the modelling itself, e.g. simulation of the complex processes of transferring heat, momentum and moisture through the atmospheric boundary layer, and secondly the starting conditions, from which the complex equations are integrated forward in time, are imperfectly specified. To put matters in perspective, modern NWP systems contain about one million points in the model global atmosphere at each of which equations are solved to produce forecast values of wind, temperatures, pressure and humidity — and yet the total number of data values available as starting conditions is one order of magnitude less than this. Thus there is a high degree of interpolation required at the start, more in some areas than others. To make matters worse the data contain errors too, and part of the analysis process is concerned with minimizing these errors — a task requiring a good deal of skill.

Despite these limitations, NWP global models consistently yield good guidance on the evolution of weather systems up to 5 days ahead, and occasionally up to 8–9 days depending on the stability of the weather situation. Most impressively the models can predict the

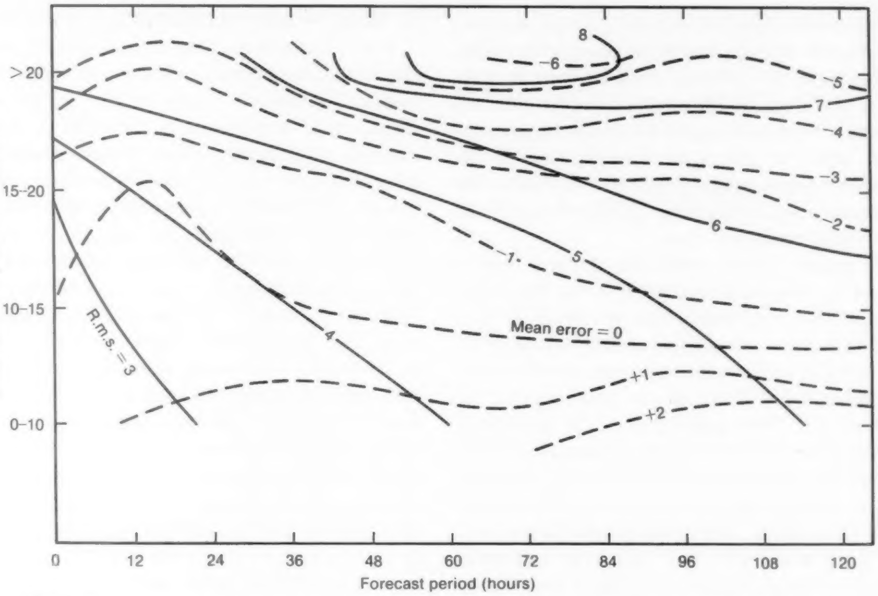

Figure 2. Errors and root-mean-square errors of wind speeds in ranges with respect to forecast period.

birth and growth to peak intensity of major depressions several days before the event. Finer-grid (but limited area) models are used to predict the evolution of sharp boundaries rather better, e.g. fronts with associated precipitation. The modelling of land topography depends upon the grid used and in order to produce realistic forecast values of weather variables for valleys, plains and mountain slopes, mesoscale models with spacing between points of only 10–15 km are used.

The above considerations lead to a broad categorization of error sources as follows:

(a) features poorly represented in the analysis,

(b) speed of development poorly simulated,

(c) speed of movement of a system too slow or too fast, and

(d) local topographical effects poorly represented.

Even small errors as defined above can lead to large inaccuracies when verification is applied to a fixed place at a fixed time. This is largely because the spatial variation of a weather element is not usually uniform; rather there are sharp gradients of wind, temperature and precipitation associated with fronts and troughs for example.

This knowledge may help the user to understand the uncertainties associated with his forecast but is there a way in which the forecaster can express this uncertainty which minimizes the range of error that the user has to allow for on a specific occasion?

5. Risk assessment — a value-adding process by the forecaster

A competent and efficient forecasting organization will have a thorough knowledge of the characteristics and formulation, and hence limitations, of the NWP models which support it. Thus systematic errors which occur in specific situations can be anticipated, e.g. a tendency to move fronts too quickly northwards in winter (say) or a tendency to underestimate the speed of development of intense depressions in certain geographical areas. By monitoring the NWP forecast and comparing with the analysis, deficiencies can be spotted early and consequences allowed for. Local forecasters develop expertise in noting how mesoscale topography can modify weather conditions in small localities. What all this means is that the forecaster can make a quantitative estimate of the likely error in a specific forecast which may be more or may be less than the average r.m.s. errors at a specific place.

It follows therefore that the forecaster could, in principle, issue modified error bands associated with each forecast, which the user may compare with the r.m.s. errors. He would then know whether to have more (or less) confidence in the forecast than usual. Unfortunately this would be a very laborious task for the forecaster but more importantly it would not necessarily help the user to make the right decisions.

To help the user interpret the (categorical) forecast it is necessary to know what values of a specific weather element are most important to him. Thus, in the example given of a forecast wind speed of 30 m s^{-1} at T+96, the crucial threshold for damage may be 35 m s^{-1} and since this figure is precisely one r.m.s. error range above the forecast then clearly it is a matter of assessing the risk. The uncertainty can be quantified by the forecaster from his consideration of all the aspects of the particular synoptic situation. If the forecaster is very confident that the wind will reach 30 m s^{-1} (perhaps with

some timing error) then the user will almost certainly decide not to risk damage and make alternative plans. On the other hand the synoptic situation may be quite volatile, e.g. different NWP models offering different solutions and solutions changing with succeeding initial conditions, and the forecaster's confidence in the 30 m s^{-1} value corresponding lower. On this basis the user may decide to delay making the decision until the situation is more certain.

It may be argued that consideration of uncertainty is best handled by direct dialogue between the forecaster and the user. This is probably true but there are two drawbacks; it may not always be possible to talk to the forecaster at short notice and, importantly, there is no objective means of evaluating the accuracy of the forecaster's advice. These difficulties can be minimized by the issue of a risk assessment table, constructed by the forecaster, which states the likelihood of a weather event occurring in a specified period. The risk is usually defined as a probability, with values ranging from 0 (no chance) to 1 (certain). Table I illustrates the concept in relation to wind speed threshold values at an offshore location. Although the scheme is best applied to specific sites for ease of verification, the principle can also be applied to areas on the assumption that a single occurrence of the event within the area implies a hit. Each user of the information should have his own thresholds to which the risk relates. On the face of it the user is now in possession of all the forecast information he needs to make his decisions — with one exception — how reliable is the forecast service? Should he make any allowance for shortcomings in the service?

One of the prime advantages of a forecast in numerical probability terms is its suitability to meaningful evaluation. There are several characteristics of forecasting performance that can be assessed:

(a) Bias — does the forecaster consistently overestimate or underestimate the risk?

(b) Accuracy — Brier scores can be calculated as a measure of accuracy.

(c) Skill — the Brier score may be compared with similar scores for climatology, persistence, NWP categorical and, of course, other forecasters.

Table I. Wind speed probability forecasts, for an oilfield, issued at 0900 UTC on 1 January 1990

Date	Period (UTC)	Wind speed (kn) probabilities >30	>45	>60
1st	1200–1800	0.4	0.1	0.0
	1800–2400	0.2	0.0	0.0
2nd	0001–0600	0.2	0.1	0.0
	0600–1200	0.6	0.5	0.3
	1200–2400	0.7	0.4	0.1
3rd	0001–1200	0.8	0.6	0.3
	1200–2400	0.7	0.4	0.2
4th	0001–1200	0.5	0.3	0.0

6. Bias — use of reliability diagrams

A lot of information about the forecaster performance can be derived from examination of the distribution of forecast probability against the observed relative frequency of occurrence of the event (summed over a number of forecast issues). Figs 3(a)–3(d) illustrate some examples. For the perfect forecaster who always issues probabilities of 0 or 1 and always succeeds there will only be two points on the graph. The realistic forecaster will use the full range of probabilities, and if he is perfectly calibrated all points will lie on the leading diagonal. In practice the forecaster is unlikely to be so good; Figs 3(c) and 3(d) show contrasting performance in the remote forecasting of rainfall (defined as wet or dry) and fog respectively for a specific location within a specified period. Apart from the curious over-optimism in use of $P = 0.9$ the forecasting of uncertainty in rainfall occurrence is quite reliable. By contrast there is clearly over-forecasting of fog which is perhaps not surprising in view of the patchy distribution of the element and the desire of the forecaster to cover his expectation in some way. (In this latter case one would expect the local forecaster to do better, given his superior knowledge of local climatology.) Reliability diagrams may be drawn up for any threshold, e.g. winds exceeding 35 m s^{-1}.

7. Skill — use of Brier scores

A probability forecast, e.g. Table I, can be verified by noting whether the event occurred or not and scored using the Brier scheme:

$$\text{score} = (V - P)^2$$

where P = forecast probability, and
V = 1 if event occurs, and
= 0 if event does not occur.

It can be seen that the best score is zero and the worst score is one.

The Brier score (BS) is objective and provides an evaluation measure that encourages forecasters to make forecasts that truly reflect their judgement, i.e. the forecaster cannot maximize his expected score by hedging. Taken in conjuction with the reliability curves, the Brier scores provide a useful quantitative measure of the quality of the forecaster's advice. Clearly one can compare the merits of different forecasting services from comparison of these parameters providing, of course, the same periods, thresholds and locations are used.

A measure of absolute accuracy is one characteristic, but skill may be quite another. In essence, skill is all about showing that the forecaster can give better advice to the user than if he simply relied upon climatology (clim) or persistence (pers) to estimate a future weather condition. This is a commercially important consideration; it is also extremely useful for the forecaster to know what he is up against! NWP and persistence forecasts are usually expressed in categorical terms, i.e.

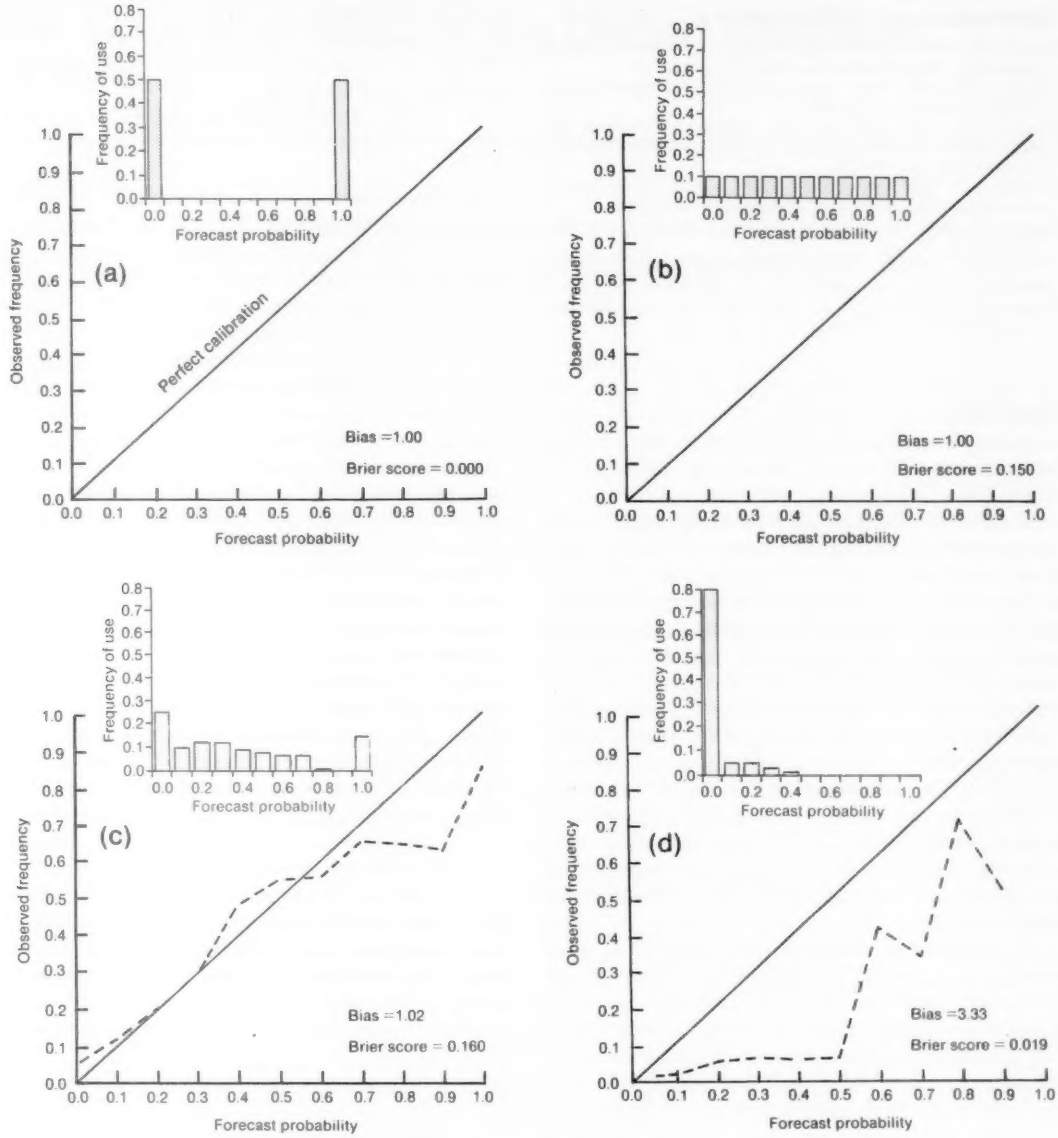

Figure 3. Forecast probabilities compared with observed frequencies for (a) perfect prognosis, (b) perfect reliability, (c) actual performance for rainfall over a period, and (d) as (c) but for fog. In each diagram a histogram of the frequency of use of the forecast probabilities is included.

probabilities of 0 or 1, in relation to specific thresholds, and can be scored accordingly. Forecasts based on climatology can be expressed in categorical or probabilistic terms, although the latter is not usually readily available. The skill is defined as follows:

$$\text{skill score} = \frac{\text{BS (clim/pers/NWP)} - \text{BS}}{\text{BS (clim/pers/NWP)}} \times 100.$$

It could be asserted that the case for probability forecasts rests on the demonstration of positive skill compared with the alternative means of expressing forecasts.

More detailed information on probability forecast evaluation is contained in the Appendix.

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The assistance of John Norris whilst a chief forecaster in the Central Forecasting Office at Bracknell, who provided the calculations of Brier scores and the reliability diagrams, is gratefully acknowledged. Also thanks to Mrs Mai McGinnigle for help with the illustrations and finally the advice and encouragement from Professor Allan Murphy was much appreciated.

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Appendix

In assessment of probability forecasts, the attributes with which we are most concerned are bias, accuracy and skill.

Overall, bias may be defined as the ratio of the average forecast probability of an event to the corresponding observed relative frequency of occurrence of that event. This ratio is variously known as global bias, bias in-the-large, reliability in-the-large, or simply bias. The last-named is used here. The optimum bias value is unity. Values less than one are indicative of under-forecasting; values over one of over-forecasting.

A further aspect of bias is the correspondence between observed relative frequency and forecast probability for subsamples of the complete data set; notably, for each of the discrete probability values available to the forecaster — in the present case the 11 values 0, 0.1,.....0.9, 1.0. This quantity is known by the names bias in-the-small, reliability in-the-small or simply reliability.

The measure of accuracy used here is the Brier probability score (*PS*) (Brier 1950). Each of the parameters examined may be considered to be a nominal binary predictand, i.e. each probability forecast is considered to be a two-event situation in which the forecast event will or will not occur. For such predictands, the Brier score may be expressed as:

$$PS = \frac{1}{N} \sum_{i=1}^{N} (V_i - P_i)^2$$

where N is the total number of forecasts, P_i is the forecast probability in the ith case, and V_i is the verifying observation. V_i takes the value 1 if the event occurs and 0 if it does not. In this form, *PS* has a range from 0 (the best possible score) to 1.

The Brier score is a 'strictly proper' scoring rule, i.e. it is an evaluation measure that encourages forecasters to make the forecast corresponding with their judgement — the forecaster cannot minimize his expected score by hedging (Murphy and Epstein 1967, Winkler and Murphy 1968).

It has been shown by Murphy (1973) that provided the forecast probabilities possess only a finite number of distinct values, then the Brier score may be partitioned into three separate terms:

$$PS = UNC + REL - RES. \qquad (A1)$$

In this expression, *UNC* represents the uncertainty and is a measure of the variance of the observations; it does not depend in any way upon the forecasts. It is, in fact, the Brier score that would result if all the forecasts consisted of the (sample) climatological frequency. Values of the *UNC* range from a maximum of 0.25 when the event occurs on half the possible occasions to a minimum of zero when the event occurs always or not at all.

REL is the reliability mentioned earlier. It is the weighted mean squared difference between the forecasts and the corresponding observed relative frequencies over all the subsamples. Reliability may be represented qualitatively by a diagram in which relative frequency is plotted against the forecast probability for each of the specific probability values. In a reliability diagram the diagonal line from the origin represents perfect, or zero, reliability.

RES is the resolution term. This is the weighted mean squared difference between the observed relative frequency in the subsamples and the overall relative frequency. It takes values between 0.25 (the best) and 0. Sanders (1963) considers that resolution reflects a forecaster's ability to 'sort' occasions into subsets.

To demonstrate forecasting skill it is necessary to compare the achieved accuracy (represented here by the Brier score) with the score that would be produced by some reference procedure. The reference procedure used here is the (constant) forecast of the overall sample relative frequency (i.e. the sample climatological probability). This is in fact given directly by the uncertainty term *UNC*.

We may thus define a skill score *SS* as follows:

$$SS = \left(\frac{UNC - BS}{UNC}\right) \times 100.$$

Substituting from (A1), we get

$$SS = \left(\frac{RES - REL}{UNC}\right) \times 100.$$

SS can take values between 100 and minus infinity. Thus, the skill of the forecasts is positive if the magnitude of the resolution term exceeds that of the reliability term. A negative skill score indicates a result inferior to that obtained by a repeated use of a constant value based upon the sample climatology.

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The autumn of 1990 in the United Kingdom

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Summary

The autumn was warm and dry over much of England and Wales but close to average elsewhere, although some places around the Moray Firth and the Firth of Forth were unusually wet; it was generally sunny except over northern and eastern Scotland, northern and eastern England and East Anglia, where it was rather dull in places.

1. The autumn as a whole

Autumn temperatures were generally above normal over England and Wales and below or near normal over Scotland and Northern Ireland, ranging from 1.1 °C above normal at places in Cambridgeshire and Dorset to 0.6 °C below normal in the west of Northern Ireland. Autumn rainfall totals were generally below normal over England and Wales and above normal over Scotland and Northern Ireland, with amounts ranging from 169% at Dunbar, Lothian Region to 52% at Lyneham, Wiltshire. Sunshine was generally above average over the autumn period, except for northern and eastern Scotland, northern and eastern England and East Anglia, where amounts were near or just below average, and ranged from 128% at Rhoose, South Glamorgan to 67% at Cockle Park, Northumberland.

Information about the temperature, rainfall and sunshine during the period from September 1990 to November 1990 is given in Fig. 1 and Table I.

2. The individual months

September. Mean monthly temperatures were generally near or below normal, ranging from 2 °C below normal in western Scotland to 0.6 °C above normal in the Channel Islands. On the 28th the temperature fell to −1.4 °C at Hurn, Dorset, giving the coldest September night there in nearly 50 years of records. Monthly rainfall totals were well above normal in northern Scotland, the Western Isles, and parts of North Wales, but below normal elsewhere, with a large part of the Midlands and south-east England having less than half the normal rainfall, and ranged from 181% at Cape Wrath, Highland Region to 36% at Folkestone, Kent and in central London. Monthly sunshine amounts were below average over East Anglia and much of Scotland, apart from the Forth–Clyde valley; elsewhere amounts were near or above average, reaching nearly 140% in South Wales. In contrast north Norfolk was rather dull with only 76% of average sunshine.

After a hot start to the month cooler weather spread from the north-west by the 3rd, bringing some rain to most places, especially in the west and north. A somewhat settled spell between the 7th and 15th was followed by generally cooler and more changeable weather. During the night of the 18th/19th heavy rain and strong winds caused a landslip on the West

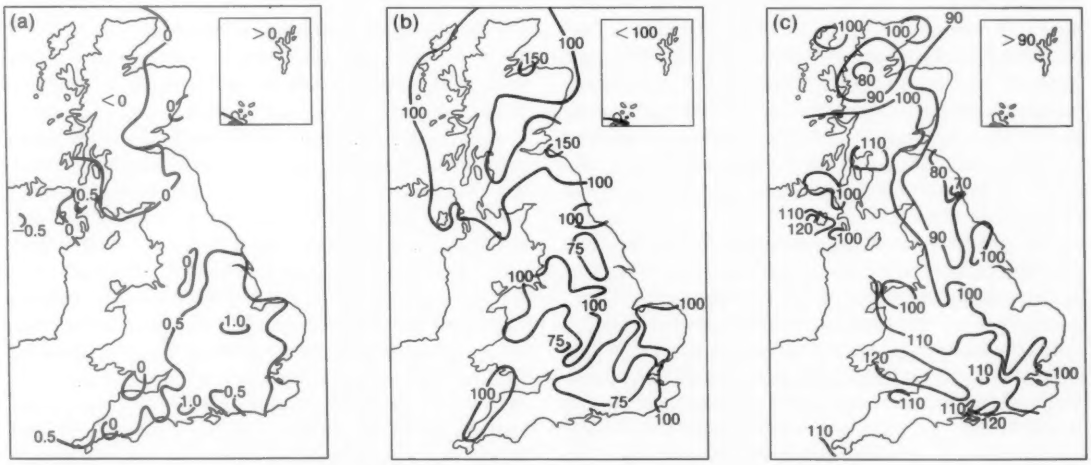

Figure 1. Values of (a) mean temperature difference (°C), (b) rainfall percentage and (c) sunshine percentage for autumn, 1990 (September–November) relative to 1951–80 averages.

Table I. District values for the period September–November 1990, relative to 1951–80 averages

District	Mean temperature (°C)	Rain-days	Rainfall	Sunshine
	Difference from average		Percentage of average	
Northern Scotland	0.0	−1	113	98
Eastern Scotland	+0.1	0	105	91
Eastern and north-east England	+0.4	0	86	89
East Anglia	+0.5	0	83	101
Midland counties	+0.6	−1	89	104
South-east and central southern England	+0.7	−1	77	114
Western Scotland	−0.1	−1	102	109
North-west England and North Wales	+0.2	−1	95	105
South-west England and South Wales	+0.4	0	87	120
Northern Ireland	+0.1	0	109	108
Scotland	0.0	−1	108	99
England and Wales	+0.5	0	87	105

Highest maximum: 26.4 °C in south-eastern and central southern England in September.
Lowest minimum: −5.9 °C in western Scotland in November.

Highland line from Fort William to Mallaig at Glenfinnan. Heavy showers on the 22nd/23rd in the south and west later became widespread over England and Wales with hail and thunder in places. It remained unsettled in most places for the rest of the month apart from a brief spell of fine weather on the 27th. Thunder was reported over Dorset during the evening of the 29th and for a time around London after noon on the 30th.

October. Mean monthly temperatures were above normal everywhere, ranging from near normal at Benbecula, Western Isles to nearly 2 °C above normal at Gatwick, West Sussex. The temperatures of 24.3 °C at Valley, Gwynedd and 22.3 °C at Kinloss, Grampian Region were the highest in their respective localities since 1959. The value of 11.8 °C in the register of Central England Temperatures indicates that October was the warmest for 12 years. Monthly rainfall amounts were above normal in many areas although some central and southern areas ended the month with less rainfall than normal. Amounts ranged from 219% at Edinburgh, Lothian Region to only 61% at Stansted, Essex. Sunshine amounts were about average over the whole of the United Kingdom, although some places were generally rather dull, and ranged from 130% over East Anglia to as little as 63% at Eskdalemuir, Dumfries and Galloway Region.

The weather was changeable over the month as a whole, being generally unsettled with one or two brighter interludes. The month was rather windy with strong to gale force winds on 8 days. In the south-east a thunderstorm on the 15th ended a very warm spell with a dramatic display of lightning reported at Colchester, Essex. Other thundery outbreaks occurred over South Wales on the 3rd, Scotland on the 16th, and mainly eastern areas of England and Wales between the 17th and 19th, and more widely spread between the 25th and 27th, sometimes accompanied by hail. During the 28th scattered thundery activity continued in southern England and South Wales, and isolated thunderstorms with hail occurred in the north-west; thunder was reported along the south coast on the 31st.

November. Mean monthly temperatures were generally about average and ranged from 0.7 °C below normal at Aspatria, Cumbria to 1.0 °C above normal at Gatwick, West Sussex. Apart from a few places in the east, mainly near the coasts, rainfall amounts were generally below normal and ranged from 30% of normal at Carlisle, Cumbria to more than 180% at Lowestoft, Suffolk. Sunshine amounts were generally above average in western and southern areas of the United Kingdom, but below average in central and eastern areas, ranging from 160% in west Cornwall to 71% around Blackpool and Manchester. Western Scotland was generally sunny in contrast to eastern areas where it was rather dull. At Paisley the sunshine total for the month was equal highest with that of 1947 in a record back to 1885, although at nearby Glasgow Airport the total was 5 hours short of the amount recorded in November 1947.

After a quiet beginning to the month the weather was unsettled between the 9th and 27th, with some rain, hail, thunder and even snow in places. Thunder was reported along the south coast on the 1st, in the Glasgow and Leeds areas on the 14th, over East Sussex on the 22nd, East Anglia on the 23rd, eastern Scotland, west Cornwall, and Kent on the 24th and Fife, the Thames Valley, eastern Kent and the Isle of Wight on the 25th, and was accompanied by hail over south-west England, South Wales and the Isle of Wight on the 21st and Cumbria and Lancashire on the 23rd. Hail was reported over North Wales on the 18th, the Isle of Man and Salisbury Plain on the 20th, Devon and Cornwall on the 23rd and northern and eastern Scotland and west Cornwall on the 25th.

Review

Remote sensing in hydrology, by E.T. Engman and R.J. Gurney. 160 mm × 240 mm, pp. xiv+225. *illus.* London, Chapman and Hall, 1991. ISBN 0 412 24450 0.

During the 1990s a number of satellite missions are planned which will provide data of great interest to hydrologists as well as meteorologists. It is timely therefore that two hydrologists should produce a book which demonstrates graphically how hydrology is expanding from the study of local or regional scales to consider the global hydrological cycle. In 11 chapters, totalling 255 pages, the reader is treated to careful descriptions, often quantitative, of the full range of hydrological subjects from basic measurements of precipitation, snow, evapotranspiration and soil moisture, through runoff and snowmelt modelling, ground water and water-quality measurements to water resources management and modelling.

This well-presented book is targetted at water resources scientists and managers, but it is likely to be of significant interest also to hydrometeorologists, and those involved in remote sensing and environmental monitoring who wish to find out more about the combined contribution of remote sensing and hydrology to their subject. We are told in chapter 1 that remote sensing offers a unique approach to studying hydrology across the disparate scales, and subsequent chapters demonstrate this. Each chapter follows roughly the same format: an introduction, general approach, measurement theory/details/examples, future applications and references.

There are some differences of emphasis and presentation between individual chapters. For example, in the description of the basic principles of remote sensing (chapter 2) there is no mathematics at all, yet in discussing snowmelt and runoff modelling there are many equations and detailed model flow charts. Similarly in some chapters we find excellent descriptions of the physics behind the measurements, a good example in chapter 7 being the description of what happens to the dielectric constant of soil as water is added to it. However in the chapter on precipitation there is no clear description of absorption and scattering processes underlying passive microwave techniques, nor the physics of ground-based radar measurement techniques.

Perhaps it is a little unfair to demand that a book on hydrology should contain such information, but at least there should be some pointers towards the source of further information. Also it would have been useful to include some comparison of techniques, particularly for measuring precipitation, and some discussion of measurement problems for various parameters. In general this book is uncritical. There is no mention also of the considerable work being undertaken to calibrate runoff models using radar data, or to use adaptive updating to cope with both measurement and forecast errors arising from the use of remotely sensed data. In addition there are a few inaccuracies which are no fault of the authors. These are due mainly to the rapid changes in plans for remotely sensing systems. For example, Table 11.1 defines the Tropical Rainfall Monitoring Mission (TRMM) incorrectly. The TRMM orbit will be at an inclination of 35°, the microwave radiometer has additional channels at 10 and 22 GHz, the radar will operate at a single frequency of 14 GHz and the launch will probably be in 1996.

Whilst these omissions and slight inaccuracies may irritate some readers, they should not be allowed to detract from the general high standard of the text. This reviewer found reading this book an enriching experience. There is no doubt that the authors are right in saying that only the microwave region of the electromagnetic spectrum offers the potential for truly quantitative measurements in many areas. Data from the ERS-1 SAR, SSM/I and TRMM present exciting opportunities for future work. Every hydrologist and hydrometeorologist should obtain this book, and read it cover to cover, as within is contained a wealth of information. This information will underpin large areas of hydrological science in the next decade. Increasingly, hydrologists will be called upon to contribute to better land surface parametrizations in global NWP models and GCMs using improved interpretation of remotely sensed data. In turn the model products and the remotely sensed data should significantly increase our understanding of continental-scale hydrological processes, and hence the impact of climate change on the hydrological cycle.

C.G. Collier

Books received

The listing of books under this heading does not preclude a review in the Meteorological Magazine at a later date.

Numerical adventures with geochemical cycles, by J.C.G. Walker (Oxford University Press, 1991. £30.00) contains a methodology for computational simulation of aspects of the earth's geochemical evolution. Programs which will execute on a personal computer are described. ISBN 0 19 504520 3.

Atmospheric data analysis, by R. Daley (Cambridge University Press, 1991. £55.00) outlines the physical and mathematical basis of all aspects of the subject, with the emphasis on theory. However, many practical considerations and examples are included, and it can be used as a textbook or reference book. ISBN 0 521 38215 7.

Satellite and radar photographs — 7 July 1991 at 0600 UTC

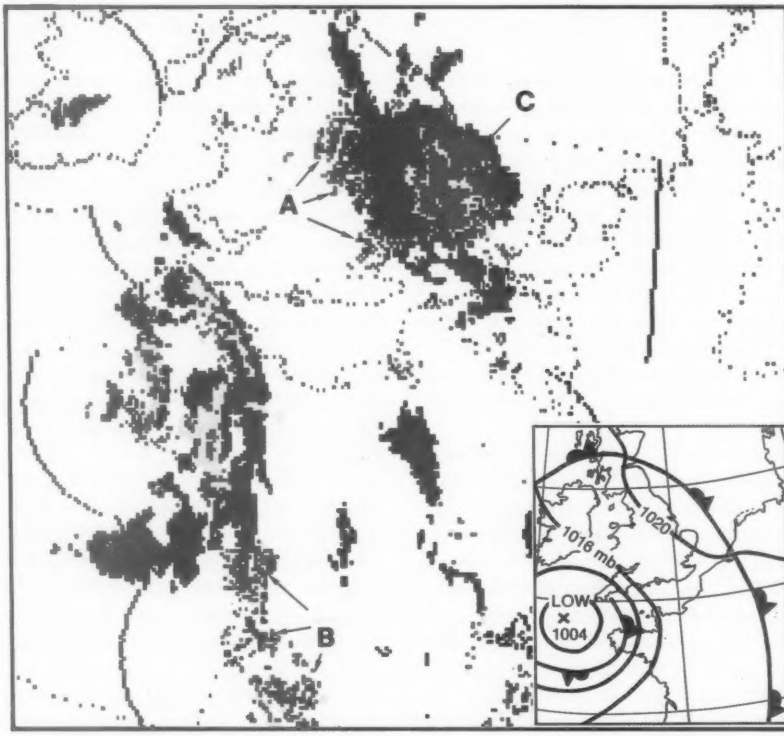

Figure 1. COST-73 satellite and radar image for 0600 UTC on 7 July 1991. Dark blue represents cloud tops with temperatures between −15 and −45 °C, and mauve <−45 °C. Rainfall intensities (mm h^{-1}) are shown as follows: green < 1, yellow 1–3, red 3–10, light blue 10–30, black > 30. Coastlines, national boundaries and the limits of radar coverage are shown in black. The surface analysis for that time is inset.

Composite satellite and radar images provide the analyst with vital information on the location and evolution of significant areas of cloud and precipitation, especially in otherwise data-sparse regions. Sometimes, however, the radar images can be misleading and apparently indicate rain in areas that are in fact dry or even cloud-free. An example is shown in Fig. 1 which was taken during a period when a warm south-easterly airstream covered the British Isles (see inset to Fig. 1). Areas of apparent heavy rain A, some collocated with the western portion of the cloud band in eastern England, are in fact returns from the ground — Anaprop. Surface reports showed these areas to be dry. Other examples can be seen in south-west France at B. Anaprop is caused by refraction of the radar beam by a low-level temperature inversion and hydrolapse (Fig. 2). Time-lapse sequences can help to discriminate these spurious echoes since they tend to remain stationary, whereas those associated with precipitation usually move. In the future, data from Doppler radars may provide a mechanism for removing these unwanted returns.

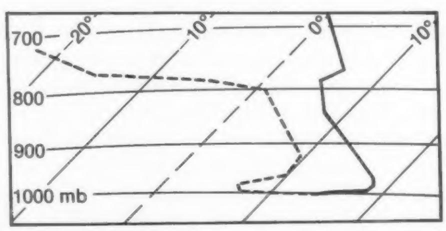

Figure 2. Tephigram for Crawley at 0000 UTC on 7 July 1991. The low-level temperature and humidity structure is representative of areas where spurious echoes have been identified at A in Fig. 1.

The large area of heavy rain at C, collocated with the area of cold cloud tops, is associated with a mesoscale convective system which developed over the southern North Sea and subsequently moved northwards. The evolution of a similar system over the near continent is shown in Lilley*.

A.J. Waters

* Lilley, R.B.E.; Satellite and radar photographs — 2 July 1991 at 1700 UTC. *Meteorol Mag*, **120**, 1991, 172.

GUIDE TO AUTHORS

Content

Articles on all aspects of meteorology are welcomed, particularly those which describe results of research in applied meteorology or the development of practical forecasting techniques.

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